A COLD LOW OVER THE FAR WESTERN STATES, AUGUST 1–5, 1956

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1. INTRODUCTION

The northeastern portion of the North Pacific Ocean and its adjacent land areas is a well known source region for the cold-type Low. Although, not unique in this type of cyclonic perturbation, this locality is one of the most prolific sources of cold Lows that affect the United States. The area of maximum occurrence tends to migrate northward along the west coast of North America during the summer months and southward in the winter. Cold Lows or "cold pools", as they are referred to by the British, have been rather well defined by Sumner [1] ". . . a mass of cold air in depth entirely surrounded by relatively warm air and appears as one or more closed lines in the thickness isopleths for any fairly deep atmospheric layer."

This article discusses the formation of a cold Low and indicates certain points that might aid in forecasting similar developments in the future. A summarization of the weather in essentially a qualitative manner is presented, as well as a few ideas that may aid in the forecasting of stability or instability and of temperature maxima or minima.

2. ANTECEDENT CONDITIONS

The well-developed subtropical anticyclone which persisted over the North Pacific Ocean during the period of interest, August 1–5, 1956, was also present for some time prior to and following this period. This High or an associated High apparently had its inception around July 15, 1956 with its center originally located from 10° to 15° of longitude east of the International Date Line. Downstream from this position a long-wave trough was established over the eastern portion of the Pacific. Coexistent with this trough and the ridge was an upstream long-wave trough over eastern Siberia; this latter trough remained practically stationary with regard to any east-west motion and was rather weak throughout most of the 3-week period, July 15–August 5.

By the 20th of July, an eastward-building ridge, extending toward the California and Mexico coasts, had developed in the southern portion of this Pacific anticyclone. Concurrently in the upper air the continental High over southeastern United States was building westward, finally coalescing with the Pacific ridge, and with this merger a new High began to appear between 145° and 155° W. Because of this development there began a redistribution in the wavelength and pressure center patterns over the Pacific Ocean and North America. The long-wave trough off the west coast weakened and moved eastward over central North America, while the long-wave ridge just east of the Date Line moved westward to near 170° E. and, for the most part, south of the 45th parallel. The Siberian long-wave trough continued in much the same position as previously, north and west of the ridge.

During the next few days a pronounced block formed, extending from the central Pacific Ocean northeastward to the vicinity of Great Bear Lake, Canada, and thence eastward across the Davis Strait and into the eastern Atlantic. By July 25 the westernmost Pacific High was weakening and merging with the high cell centered at 40° N., 145° W. The Siberian long-wave trough had intensified, and this was to play an important part in future downstream developments through the dispersion of energy. The block, which then extended from the mid-Pacific area to Alaska and thence across northern Canada, had intensified during the past day or two. This intensification increased the downstream transport of cold Arctic air west-southwestward into southern Canada thus producing and developing a trough from Hudson Bay to British Columbia. The chart (not shown) of the 700-mb. 5-day mean height departure from normal for the 5-day period centered on July 26 indicated the following positive anomaly center values: 420 ft. at 50° N., 165° W; 450 ft. approximately 330 miles north of Great Bear Lake; and 370 ft. midway between Labrador and southern Greenland. Negative anomaly center values were 190 ft. near Wrangel Island, 200 ft. at the North Pole, and 230 ft. over southern Hudson Bay. The Hovmöller Chart [2] in figure 1 shows that also on this date, July 26, the 500-mb. trough associated with the Siberian Low reached its maximum intensity.

Between July 26 and 28, 1956, both at the surface and aloft, the anticyclone over the Pacific retrograded with its central position becoming nearly stationary at 50° N. and 165°–170° W. During this regression a series of surface waves and attendant short-wave troughs caught in a strong meridional flow around the western periphery of the large Pacific anticyclone moved northward toward the

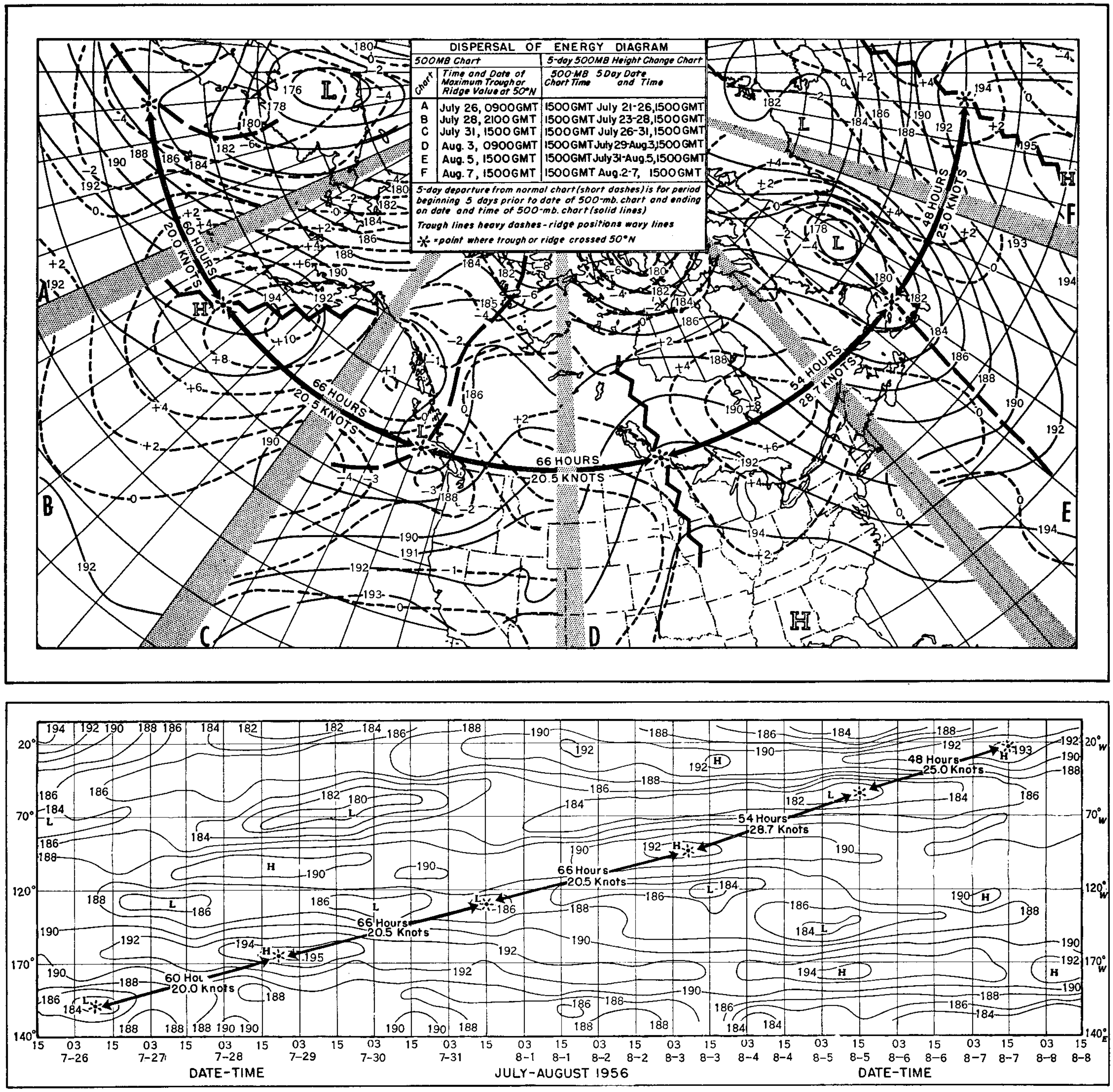

FIGURE 1—Upper—Composite 500-mb. chart in six sectors illustrating the period and time of maximum intensity of troughs and ridges as a result of downstream dispersion of energy at 50° N. Lat. Heights are in hundreds of feet. Dashed lines represent the 5-day change of 500-mb. height for the period immediately prior to the time of the 500-mb. chart. Positions at maximum intensity are clearly marked by heavy dashed lines for troughs and heavy zig-zag lines for ridges. Speed of dispersion and time in hours between periods of amplification are placed between arrows. Lower—Hovmöller diagram presenting a time-longitude cross section of the 500-mb. height at 50° N., extending from 150° E. eastward to 5° W. during the period 1500 GMT, July 25, 1956 to 1500 GMT, August 8, 1956. The heavy black line connecting alternate ridge and trough centers represents the trajectory of the dispersion of energy downstream the same as indicated on the upper chart by the arrows along the 50° N. parallel. Heights are in hundreds of feet. The H and L do not normally locate the center of an anticyclone or cyclone but merely indicate the highest or lowest value attained in ridges or troughs at 50° N.

region of the Pole. As these waves moved into the region of the cyclonic vortex there occurred a continued lowering of the central height value in the long-wave trough which originally had been located near Wrangel Island but was then being carried northward toward the Pole. Also occurring during the recession of the Pacific High center was the re-establishment of the eastern Pacific trough. This was occasioned by the depression over southern Canada building westward across the southwestern Provinces and thence southward along the Pacific Coastal States. By July 29 this southward extension of the eastern Pacific trough had severed the east-west ridge joining the Pacific and the continental Highs.

The Siberian Low, by July 29, had assumed a more circular appearance and had intensified with the central value of the 5-day mean 700-mb. height departure from normal having lowered 300 ft. since July 26. During this same period the central value of the Pacific High had risen 100 ft. in height. This isallobaric intensification and the weakening of the long-wave trough to the northwest of the High produced a westward displacement of the ridge line. Thus, there was established a strong isallobaric gradient between these two pressure systems that attained a value of 600 ft. between 65° N. and 75° N. on the chart of 5-day mean 700-mb. height departure from normal. A similar but more intense pressure gradient was indicated by the daily 700-mb. and 500-mb. charts with air flow becoming more zonal than meridional from north central Siberia into Alaska and the Yukon. Under this air flow cold Arctic air was rapidly transported into Alaska and the Yukon, gradually but definitely weakening the persistent ridge that had prevailed for some time across Alaska and northern Canada. Also, recurrent short waves, attending surface waves which were moving downstream in this westerly flow, continued to erode the ridge, and by the 31st of July a zonal flow had been established across northern sections of British Columbia and Alberta. Concurrent with this development was the formation of a cut-off cold Low with a central height of 18,500 ft. at 500 mb. near Port Hardy, Vancouver Island, B. C. Graham [3] found in his studies covering the years 1947–52 that this locality was most prolific of cyclones during the summer season reaching a percentage frequency of 2.2 for the occurrence of cyclones at 500 mb. per 100,000 km.^{2}. This value, by the way, was the highest noted for any season of the year within the area from the west coast of North America to 145° W. and from 25° N. to 55° N. Furthermore, in this instance the downstream dispersion of energy was approaching its maximum intensity along the 50th parallel between 125° and 130° W., furnishing additional support for the development of the closed Low. The long-wave trough, within which this cold Low was enclosed, extended southwestward from Port Hardy, Vancouver Island to southeast of Johnston Island.

3. DEVELOPMENT OF THE COLD LOW

The development of the cold Low near Port Hardy at the end of July 1956 occurred during a seclusion process in the upper air. Originally the Low was a portion of a cold trough that had formed on the downstream side of a pronounced high-latitude block which extended from the mid-Pacific northward and eastward along the southern limits of the Arctic Circle. The trough in this particular case had a southwest-northeast orientation and reached from central Canada to southern British Columbia and thence into the Pacific. A jet stream was located along the western periphery of the Pacific ridge and in the levels beneath the jet warm air was being transported northward and eastward. But with the tightening of the gradient along the northern portion of the ridge and the steepening of the thermal gradient between the Low centered north of Wrangel Island and the Pacific High center, a resultant acceleration in the speed of the jet stream occurred. Thus, by July 29–30, as the jet stream recurved southeastward around the northern portion of the Pacific ridge, the increased eastward momentum resulted in the transporting of warmer air slowly across height lines. Gradually, due to this eastward transport of warm air and also to subsidence as indicated by negative vertical motion values on the Joint Numerical Weather Prediction charts (not shown), there occurred a seclusion of the cold trough. This pattern of formation of the cold Low followed much the same development as presented by Palmén and Nagler [4] who stated: "Thus the polar air in the cyclonic area to the south was completely secluded from its original source region . . ." And in the same manner by July 31, 1956, the cyclonic system that is the cold Low studied in this article had been born.

4. SYNOPTIC CONDITIONS, AUGUST 1–5

The surface and upper air charts for August 1, at 1230 and 1500 GMT respectively (fig. 2) indicated a definite southward movement of the cold Low, which on the previous day had formed near Port Hardy, B. C., to near Astoria, Oreg. Considerable intensification had occurred during the past 24 hours with the innermost closed height line at 500 mb. being 18,400 ft. This height value represented a negative departure in excess of 500 ft. from the normal 500-mb. height at the anomaly center and a change of −400 ft. during the past day. It is difficult to associate this cut-off Low with any of the surface depressions other than possibly the weak cyclonic center near Yakima, Wash. During the past day the 1000–500-mb. thickness packing northwest of the stationary surface Low near Boise, Idaho, had intensified considerably. Because of this it became necessary to extend the stationary front southwest of Boise and to change its classification to a cold front of strong intensity. That cold air was moving southeastward was

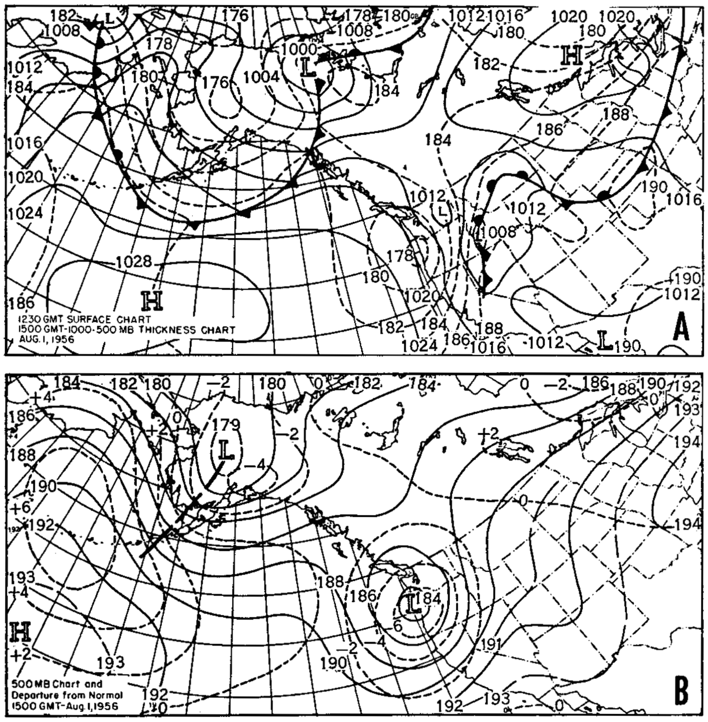

FIGURE 2.—Synoptic patterns for August 1, 1956. (A) 1230 GMT surface chart (solid lines) with 1000-500-mb. thickness for 1500 GMT (dashed lines). Surface reflection from upper cold Low appears extremely weak over eastern Washington State while intensity of gradient of thickness over western Oregon clearly defines pool of extremely cold air. (B) 1500 GMT 500-mb. chart shown by solid lines. Departures from the normal 500-mb. height for August are shown by dashed lines. Strong isallobaric departure from normal height is practically symmetrical with cold Low.

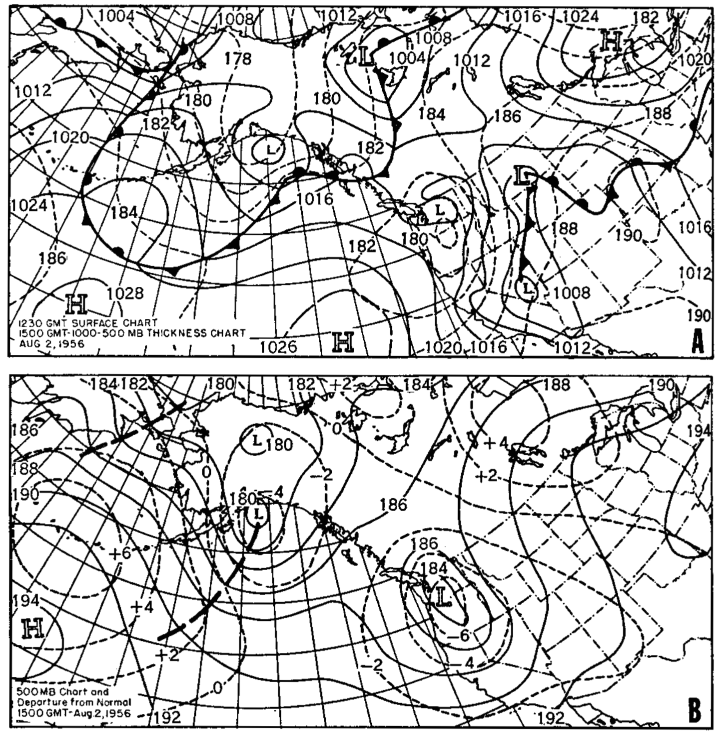

FIGURE 3.—Synoptic patterns for August 2, 1956. (A) 1230 GMT surface chart (solid lines) with 1000-500-mb. thickness for 1500 GMT (dashed lines). Surface reflection from upper cold low continues weak; little change in thermal gradient. (B) 1500 GMT 500-mb. chart shown by solid lines. Departures from normal 500-mb. height for August are shown by dashed lines. Low center practically stationary with no change in intensity.

borne out by the 24-hour thickness change chart which indicated height falls of 500 or 600 ft. at Medford and Salem, Oreg. As the mean temperature of the 1000–500-mb. layer approximates the 700-mb. thermal field, this change depicts temperature changes between 7° and 9° C. at those two stations in the past day.

In the north central and northwestern corner of figure 2A, there may be observed two of the minor waves that were moving about the intense Low north of Siberia. The Low near Great Bear Lake, Canada, although relatively weak as was the attendant cold front, was associated with a well-developed trough upstream at 500 mb. and an area of maximum cyclonic vorticity.

These two Lows, one near Great Bear Lake and the other the cold Low, produced a rather well-formed col or deformation field over British Columbia, Alberta, and Saskatchewan, Canada. From streamline flow it appeared that, in general, this could be classed as deformation with negative rotation except for slight asymmetry in the shape of the hyperbolas due to slight divergence. Thus, this would be a case in which negative vorticity was present and so any cyclone that might be approaching this col area would tend to slow down. The pair of high pressure cells that were associated with this col were the two persistent long-wave ridges that had been over the central or eastern sections of the United States and the old mid-Pacific subtropical anticyclone. A casual glance at the 500-mb. height departures from normal clearly illustrates the current intensity of the Pacific anticyclone with heights in the Bering Sea locally in excess of 600 ft. above normal, while over the central portion of the United States heights in the ridge were only slightly above the monthly average.

The warmth of the upper air in the Bering Sea and northern Canada is shown quite well on the 1000–500-mb. thickness charts by the 18,400 ft. thickness line being far to the north of its normal position and by the 500-mb. and 700-mb. temperatures which were 5° to 7° C. above normal for these regions. On the other hand, temperatures in the central portions of the cold pools, as delineated by the low thickness values over Oregon and Alaska, were indicated by the departures from normal thickness (not shown) for these areas to be below the normal for August by 6° to 9° C. at 700 mb. and by 5° to 7° C. at 500 mb. The areas of maximum 24-hour 500-mb. height rises or falls during the past day were quite well aligned with the ridges and troughs. These height changes showed the ridges had built to the north and east, while

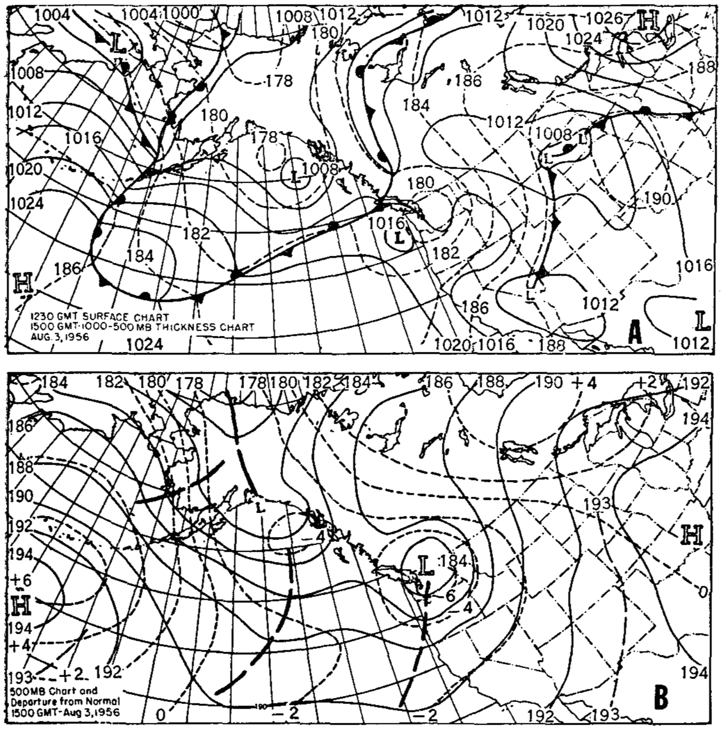

Figure 4.—Synoptic patterns for August 3, 1956. (A) 1230 GMT surface chart (solid lines) with 1000-500-mb. thickness for 1500 GMT (dashed lines). (B) 1500 GMT 500-mb. chart shown in solid lines with departures from normal 500-mb. height for August in dashed lines. Continued southward weakening of height values over eastern Pacific clearly indicated.

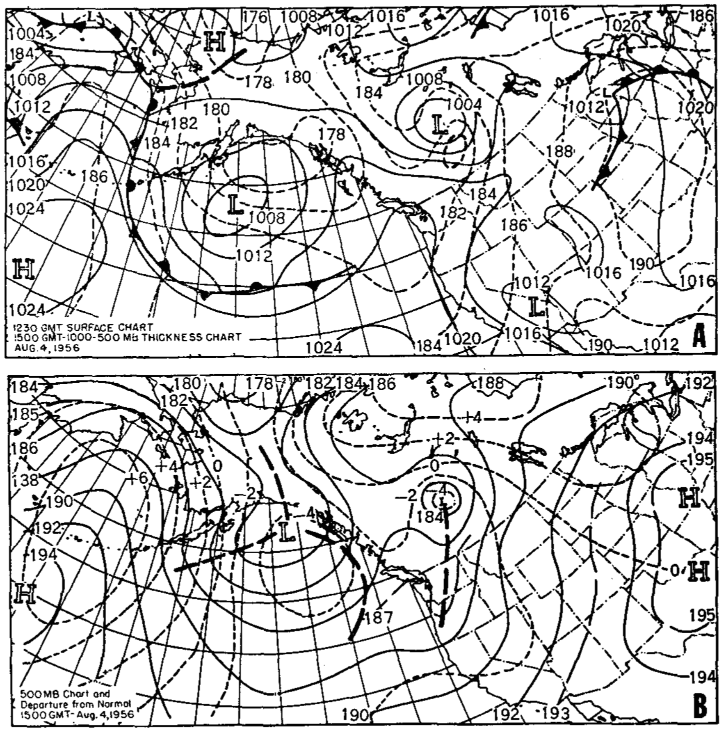

Figure 5.—Synoptic patterns for August 4, 1956. (A) 1230 GMT surface chart (solid lines) with 1000-500-mb. thickness for 1500 GMT (dashed lines). (B) 1500 GMT 500-mb. chart (solid lines) with departures from normal height for August in dashed lines. Building of ridge northward toward Alaska is plainly visible.

the troughs had deepened to the south and southeast. Height rises of 400 ft. had occurred in the northern Bering Sea and western Hudson Bay, while falls of similar magnitude were near the negative anomaly centers on the 500-mb. chart.

On August 2, 1956 the 1200 GMT surface and the 1500 GMT 500-mb. charts (fig. 3), indicated only minor changes in the intensity of pressure centers and in the position of the ridges or high centers, and nearly normal movements of fronts and troughs. The few changes may be sketched briefly. The cut-off Low moved inland over Washington State where it was centered near Yakima. A new cyclonic development appeared south of Anchorage, Alaska as the area of maximum vorticity moved into that region and a minor wave formed south of the cyclonic development. The large Pacific High which previously had encompassed most of the North Pacific had developed a second center about midway between Hawaii and the Puget Sound region. South and southwesterly winds about the western portion of these anticyclones had continued to transport warm air northward and eastward of the high center, thus permitting a continued building of the ridges. This was most pronounced over central Canada where the chart of the 500-mb. height departure from normal showed a 400-ft. positive anomaly, an increase of 200 to 300 ft. from the preceding day. Most other anomaly values of height and thickness were relatively constant even though another minor trough had appeared over eastern Siberia on the 500-mb. chart. The area of deformation with negative rotation continued to be maintained over southwestern Canada.

By August 3 the surface center of the cold Low (fig. 4A) was still poorly defined. At 500 mb. (fig. 4B) the area enclosed by the 18,400-ft. height line had increased as a trough developed between this center and the Gulf of Alaska Low. Thickness changes during the past day indicated slight modification of the temperatures associated with the cold Low over Washington State as well as a slight drift of the cold air to the north-northeast. Over the eastern Pacific the gradual weakening of the high cell, both surface and aloft, continued as cooler air was transported southward to the east of Hawaii. However, concomitantly with this weakening, the ridges over the Bering Sea and central Canada continued unchanged or built slightly. This building was most pronounced in the Canadian region for it was about this time that the eastward progression of amplitude increases in troughs and ridges reached a maximum in the ridge at 50° N., 95° W. (fig. 1). Over Siberia continued advection of warm air northeastward toward Alaska was indicated, with eastern Siberia reporting 700-mb. temperatures as much as 9° C. above normal. Minor waves continued to move southward and eastward along the western periphery of the long-wave trough that extended from the States of Wash-

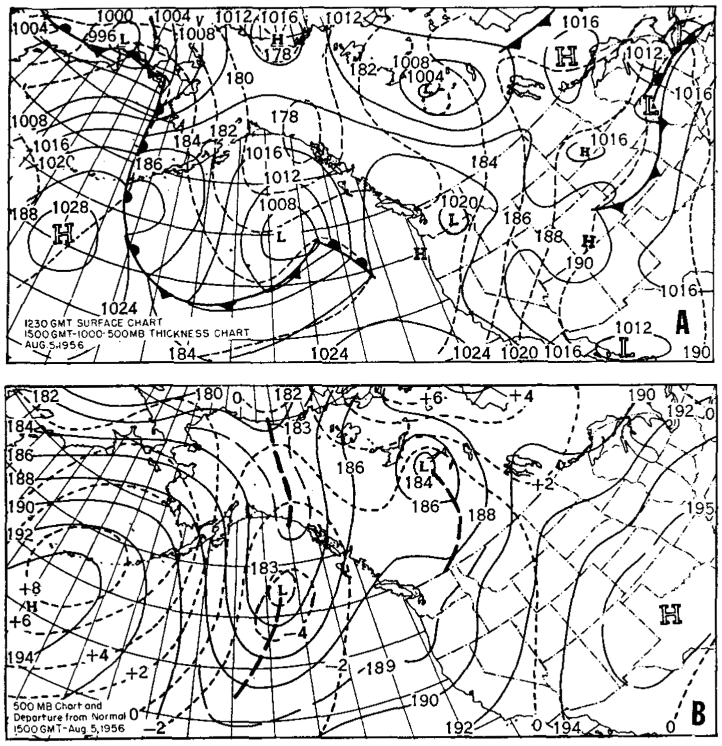

FIGURE 6.—Synoptic patterns for August 5, 1956. (A) 1230 GMT surface chart (solid lines) with 1000-500-mb. thickness for 1500 GMT (dashed lines). Thickness packing about cold Low practically dissipated as Low moved northward to more comparable temperature values. (B) 1500 GMT 500-mb. chart (solid lines) with departures from normal height for August in dashed lines. Weakening of trough over Alaska indicated as ridges continue to strengthen.

ington and Oregon northwestward to the intense cyclone centered north of the Bering Strait. The col at this time appeared to be assuming a more neutral character or, after Petterssen [5, 6], the hyperbolic streamlines tended to indicate a region of nearly pure deformation or an area without vorticity.

The surface and upper air charts for 1230 and 1500 GMT of August 4, (fig. 5) indicated, for the first time, a well formed and developed surface center in conjunction with the cold Low. The col had assumed a deformation pattern with positive rotation or cyclonic vorticity. In line with Petterssen's [5] statement that "Cols of cyclonic vorticity represent a potential cyclogenesis," it is believed that this cyclonic vorticity was as least partially responsible for the intensification of the surface Low. Along with the surface intensification there continued a slow modification of the low temperatures aloft and slight weakening of the upper centers. Temperatures at 700 mb. were from 3° to 5°. C. lower than normal in the region of the cold Low and attendant trough.

Continued slow southward transport of cool air had been maintained over the Pacific from Hawaii to the mainland with a resultant gradual decrease in heights over that region. The Gulf of Alaska cold Low, although increasing in area, continued to remain stationary aloft while drifting slowly southwestward at the surface.

Meanwhile during the past day there had occurred a rather rapid north-northeastward movement of the cold Low from Washington State to near Edmonton, Alta. both at the surface and aloft. This northward motion was perhaps due in part to the cessation of large-scale warm air transport aloft over Canada and the increased transport of cold air into the area north of the Low. Other possible explanations for the northward motion are that Lows tend to move counter-clockwise about each other, the "Fujiwhara" effect [7], and also that with the continued existence of the strong anticyclone over southeastern Canada and the United States any eastward motion was effectively impossible. Furthermore, heights to the north were lowering, as indicated by the decrease in the breadth of the ridge area. Concomitant with these occurrences was the continued building of the ridge, at the surface and aloft, over the Bering Sea and northward into extreme eastern Siberia and western Alaska due to the persistent transport of warm air into that section. By now temperatures over a goodly portion of eastern Siberia were from 9° to 11° C. above normal at 700 mb. as indicated by the 1000–500-mb. thickness departure from normal. Ahead of this building ridge was observed a short-wave trough with an intensifying zone of cylconic vorticity which was moving southeastward into a weak frontal region. Meanwhile the short-wave trough approaching the west coast had weakened considerably during the past 24 hours.

By August 5, 1956 (fig. 6) the effect of the cold Low over the Northwestern States was diminishing rapidly, but a rather persistent trough remained across the States of Washington and Oregon while the Low moved nearly northward to Fort Smith, Alta. Little change occurred insofar as the intensity of the actual 500-mb. height value was concerned but the packing of the 1000–500-mb. thickness diminished as temperatures at 700 mb. and 500 mb. returned to near normal. This change indicated a tendency for the Low to weaken henceforth, since its northward movement had removed the thermal field discontinuity, leaving it associated with air of almost uniform temperature. At the same time the trough across eastern Alaska had weakened and the upper-air height values had returned to near normal to the west of Aklavik, Canada. Over the Pacific the High began to intensify and to increase in size as a strong zonal flow along its northern periphery began noticeably to transport warmer air well into Alaska and the western part of the Gulf of Alaska. The cold Low previously in the Gulf was now nearing ship "Papa" (50° N., 145° W.) with slight weakening of the central value, but intensification had occurred in the trough that extended southward. However, this change was short-lived as during the next two days continued eastward transport of warm air dissipated the cold trough over the Gulf of Alaska, and by the 8th of August above normal heights had bridged from the Alaskan Gulf to the Washington coast. With this change in the pressure pattern, the polar troughs were deflected to the east of the Continental Divide.

5. DISPERSION OF ENERGY

Before the discussion of the relationship of energy dispersion to occurrences of trough or ridge amplification, as analyzed in this study, a brief review of this subject may be in order. Rossby [8] states:

> Practically all forms of wave motion encountered in the atmosphere or in the ocean are dispersive, i. e., the speed of propagation (phase velocity) of individual waves is dependent on their wave length. In such wave systems energy is propagated at a speed which normally differs from the phase velocity. If the speed of propagation of energy exceeds the phase velocity, new waves will be generated ahead of the initial wave train.

Later Carlin [9] stated,

> If two sets of waves of slightly different wave length are traveling through a medium and the velocity of the waves (phase velocity) is a function of the wave length, then one of the sets of waves will travel faster than the other . . . The velocity of the regions of reenforcement or interference is called the *group velocity* . . . The increase and subsequent decrease in wave amplitude is often observed to travel along a wave train at a rate much greater than the phase velocity of the wave, and it is this phenomenon which is believed to be directly associated with group velocity.

And more recently Petterssen [6] gave a very lucid discussion of the subject. He states that the group velocity, or the downstream speed of the envelope wave is greater than the zonal wind speed, and that the speed of downstream propagation of the amplification of the individual troughs and ridges gives the magnitude of the group velocity. The speed and intensity of the amplification is well illustrated by and usually extrapolated from a Hovmöller [2] diagram. Petterssen further points out that the very long waves, when identified by either space smoothing (e. g., as by the space mean chart) or by time smoothing (e. g., the 5-day mean chart as used by the Extended Forecast Section), usually remain nearly stationary for rather long periods, while the individual troughs and wedges advance through the long-wave trend at an approximate speed of 10° long. per day; on the other hand, the group velocity or amplification of the individual troughs and ridges has a downstream propagation speed of approximately 30° long. per day.

As has previously been mentioned, one of the contributory factors in the intensification of the trough over southwestern Canada and the eastern Pacific and the subsequent development of the cold Low over British Columbia is believed to have been the dispersion of energy downstream in the mid-tropospheric long-wave pattern. This amplification of the individual troughs and ridges was found to have propagated downstream with a group velocity of the order of 20 knots during the past 5 days. The initial appearance of an area of reinforcement was noted on the Hovmöller diagram (fig. 1–lower) at latitude 50° N., 164° E. upon the intensification of a trough at that location. The 24-hour 500-mb. height changes failed to show any conclusive indications of that intensification, but upon preparation of a 5-day 500-mb. height-change chart, a fall area of moderately large proportions for summertime conditions was observed with a central value slightly in excess of −700 ft. (fig. 1–upper, sector A). This maximum value occurred on July 26, at 0900 GMT. In a case study of dispersion of energy Carlin [9] stated that 5-day charts were more satisfactory than daily charts in indicating height changes and planetary movement.

Some 60 hours later (July 28, 2100 GMT) the Hovmöller diagram (fig. 1–lower) indicated definite intensification of a ridge at 50° N., 165° W. with the 500-mb. height change for the 5 days prior to this time indicating a maximum reinforcement slightly in excess of 1,000 ft. This downstream propagation of amplification had traveled at a speed of approximately 20 knots, while the computed zonal wind speed was 5° of long. per day or nearly 14 knots.

The next stage of trough amplification occurred along the west coast of North America at 50° N., 130° W. in association with the cold Low development (fig. 1–upper, sector C). In this region the picture of true deepening was masked somewhat by the presence here 5 days prior to this period of a trough which dissipated and now had reformed and was intensifying; thus, the overall change appeared less pronounced than it was actually. Proof of this was indicated by the height departures from normal in excess of 500 ft. in that region within a few hours of the 1500 GMT observations. The downstream movement from wave trough to wave trough had occurred during the preceding 120 hours at a speed slightly above 20 knots while the zonal wind speed for this area during the same period averaged approximately 11 knots, thus indicating it definitely was not phase velocity operating; nor were there any individual troughs or ridges propagating eastward through this region.

By 0900 GMT of August 3, this downstream dispersion of energy had fortified the ridge, as indicated on the Hovmöller diagram at 50° N., 95° W., with the 5-day central height rise of 800 ft. (fig. 1–upper, sector D and fig. 1–lower). This ridge formed a positive barrier against eastward motion of the Low and thus impeded its movement from the States of Washington and Oregon. During this 66-hour period the propagation of amplification or group velocity was 20.5 knots.

From figure 1 upper, sectors E and F, and figure 1 lower, it is readily observed that the downstream dispersion of energy continued to progress eastward along latitude 50° N. However, it should be noted that the speed of increase and subsequent decrease in wave amplitude accelerated considerably during the final few days of the series, the group velocity reaching 28.7 knots during one of these periods.

TABLE 1.—*Speed of downstream dispersion of energy as related to the time of maximum amplification of troughs and ridges*

Time of maximum intensity (GMT)	Length of period (hours)	Zonal index (kt.)	Group velocity (kt.)		Phase velocity (kt.)
			Computed	Observed	
0900, July 26	60	13.8	22.5	20.0	5
2100, July 28	66	8.0	19.5	20.5	−3.5
1500, July 31	66	13.6	21.0	20.5	−6.7
0900, August 3	54	13.7	27.5	28.7	0
1500, August 5	48	16.0	28.4	25.0	−3.6

Table 1 is based on the time and place of progressive amplification propagated downstream as observed from the troughs and ridges as represented by the Hovmöller chart and the composite 500-mb. chart (fig. 1). The rate of energy dispersion or group velocity is computed by dividing the linear distance between places of amplification of troughs and ridges by the time interval between their occurrences. This table also presents the computed values for the dispersion of energy downstream by use of the formula derived by Rossby [8]:

$$C_g=2U-C$$

where U represents the zonal speed of the westerlies, C the phase velocity of the planetary wave, and C_g the group velocity. The values of C_g calculated from the zonal wind speed value for 50° N., as computed twice daily by the U. S. Air Force, were compared with the observed progression or retrogression of the troughs and ridges as determined from the 5-day 700-mb. chart prepared by the Extended Forecast Section; the computed and observed values were strikingly similar in practically every case.

These obtained values are in excellent agreement with the findings of Klein [10] for the month of August 1953 when the observed speed averaged near 18 knots for downstream dispersion of energy. Carlin [9] observed and computed speeds of approximately 40 knots or 16° of lat. per day during the winter months, while McQueen and Shellum [11] observed speeds of 26 knots in June 1956. Since the rate of energy dispersion is directly dependent upon the zonal index, and as the zonal index decreases during the summer months, it appears reasonable to expect decreasing values in the speed of downstream dispersion during the summer months.

6. STABILITY OF THE AIR MASS

The instability of any cold Low may easily be deduced by observing the cloud types associated with Lows of this classification. However, it was desired in this study to indicate in some quantitative degree the stability or instability accompanying this cold pool. For this purpose the Showalter Stability Index [12] chart prepared twice daily by NWAC was first investigated. This well-known chart depicts the pattern of convective instability, with the +4° C. line representing the approximate border line between stability and instability. The larger the number is algebraically, the greater the stability and the smaller the number, the greater the instability. These charts for the period under study indicated that most of the precipitation area was located within the unstable area. Portions of these charts are illustrated by figures 7 and 8.

However, not completely satisfied with these diagrams of stability, we turned to a new series of investigations by Showalter [13] "The Stability and Thickness Relationships of Two Adjacent Layers." This is a means for easy and rapid presentation of the stability factor and by use of it one can quickly reveal the area of conditional instability. It is thought that a discussion of this relationship and how it is obtained would be beneficial before proceeding with the study.

"The Stability and Thickness Relationship of Two Adjacent Layers," as derived by Showalter comprises the use of thickness charts or individual thickness values, the number of these employed being dependent upon the height to which one desires to compute the stability of the different layers as well as the size of the area covered if individual thickness values are employed. In NWAC the following symbol system is applied:

$Z_{8.5}$ Thickness of the 1000–850-mb. layer
Z_7 Thickness of the 1000–700-mb. layer
Z_5 Thickness of the 1000–500-mb. layer
H_0 Height of the 1000-mb. surface
H_7 Height of the 700-mb. surface
H_5 Height of the 500-mb. surface

plus similar symbols for other heights and layers. (Henceforth in this article the symbols will be used for the sake of brevity.)

In 1953 Vederman and Smith [14] derived the empirical relationship:

$$H_7=\left(\frac{H_0+H_5}{2}\right)+F$$

for the derivation of the H_7 chart from the H_5 chart. Using summer data they arrived at a value of 600 gpft. for the term F. Subsequent statistical and theoretical investigations have revealed that a median value of F, adequate both geographically and seasonally, would be 500 gpft.

Interest in considering thickness ratios for other combinations of layers led Showalter to a more formal and complete evaluation of this Z_7, Z_5 relationship (fig. 9) and the formulation of the following equation:

$$Z_7-Z_5/2=+0.05\ Z_7\ \pm0.0212Z_7$$

While 0.05 Z_7 may have a range of values from +405 gpft. to +515 gpft. when the range of the 1000–700-mb. thickness chart is considered, nonetheless the overall median value of +500 gpft. is considered as adequate for practical analysis. The extreme values obtained from $\pm0.0212\ Z_7$ represent the range to be expected from appli-

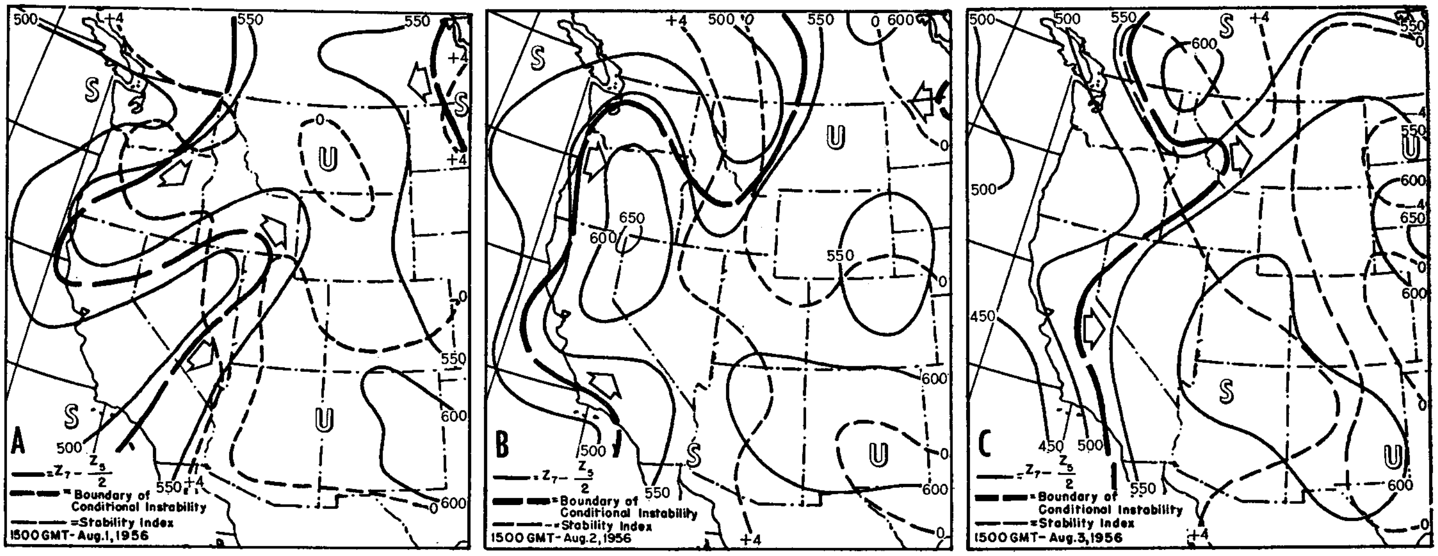

FIGURE 7.—Synoptic patterns of the stability and thickness relationships of the 1000-700-mb. and 1000-500-mb. layers, August 1–3, 1956. The solid lines indicate values of the quantity $(Z_7-Z_5/2)$ in geopotential feet at 50-ft. intervals. Values below 500-ft. line indicate more stability and values above a trend to greater instability, the end limits being near 250 and 700 feet, respectively. The dashed lines represent the Showalter stability index values in units of 4° C. Values of +4° C. or less tend toward convective instability; those above +4° C. are generally stable. The heavy dashed lines are the limits of conditional instability with the arrows indicating the direction of increasing instability. (A) 1500 GMT, August 1. (B) 1500 GMT, August 2. (C) 1500 GMT, August 3, 1956.

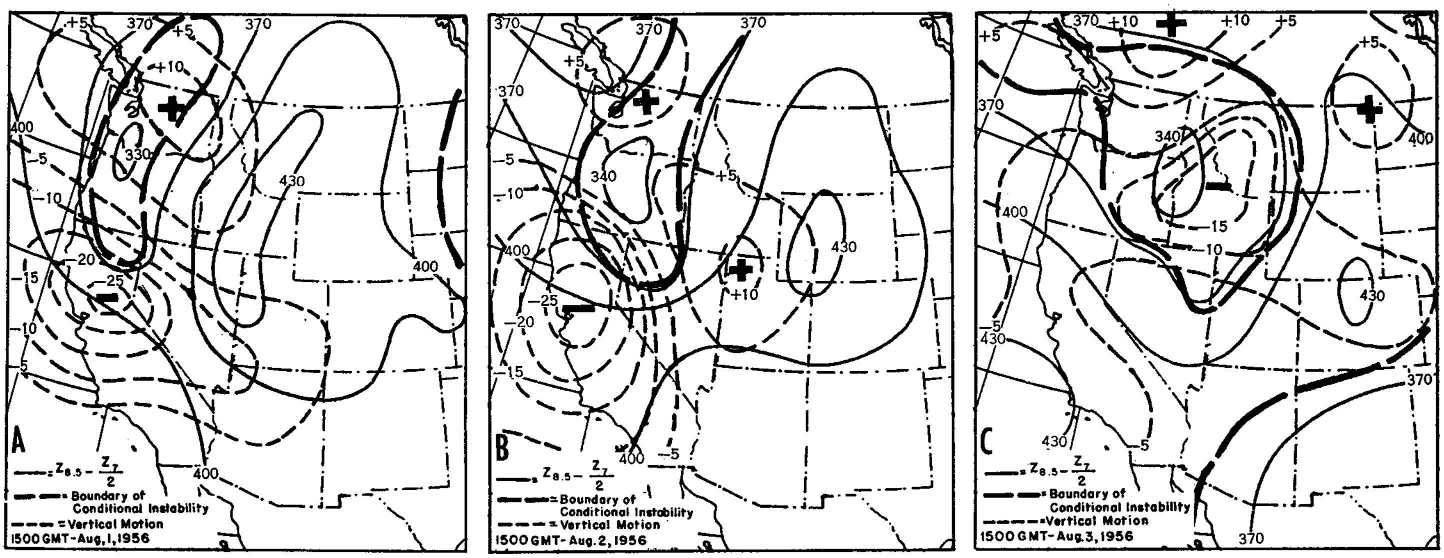

FIGURE 8.—Synoptic patterns of the stability and thickness relationships of the 1000-850-mb. and 1000-700-mb. layers, August 1–3, 1956. The solid lines indicate values of the quantity $(Z_{8.5}-Z_7/2)$ in geopotential feet at 50-ft. intervals. These values are negative and the mean value between stability and instability is near −400 gpft. Increasing departures from −400 gpft. toward positive numbers represent increasing instability and increasing departures toward more negative values represent increasing stability. The end values are near −250 and −550 feet. The heavy dashed line encloses the area of conditional instability which is always toward smaller negative numbers. Vertical motion (mm. sec.$^{-1}$) as computed by JNWP is shown by short dashed lines with rising vertical motion indicated by a plus sign and sinking or subsiding motion by a minus sign. (A) 1500 GMT, August 1. (B) 1500 GMT, August 2. (C) 1500 GMT, August 3, 1956.

cation of the formula to a stable or unstable lapse rate. And so values of $Z_7-Z_5/2$ greater than 500 gpft. are indicative of a trend toward instability while values less than 500 gpft. represent greater stability. The extreme values of stability for the $Z_7-Z_5/2$ relationship approximate 750 gpft., for which value the lapse rate of the sounding coincides with the dry adiabat, and 250 gpft., for which the lapse rate of the sounding is approximately isothermal.

The actual stability value of these two adjacent layers is easily obtainable from Z_7 and Z_5 charts by graphical subtraction techniques which permit values for an area the size of facsimile section 1 to be computed in a few minutes. However, if only a small area is desired for the

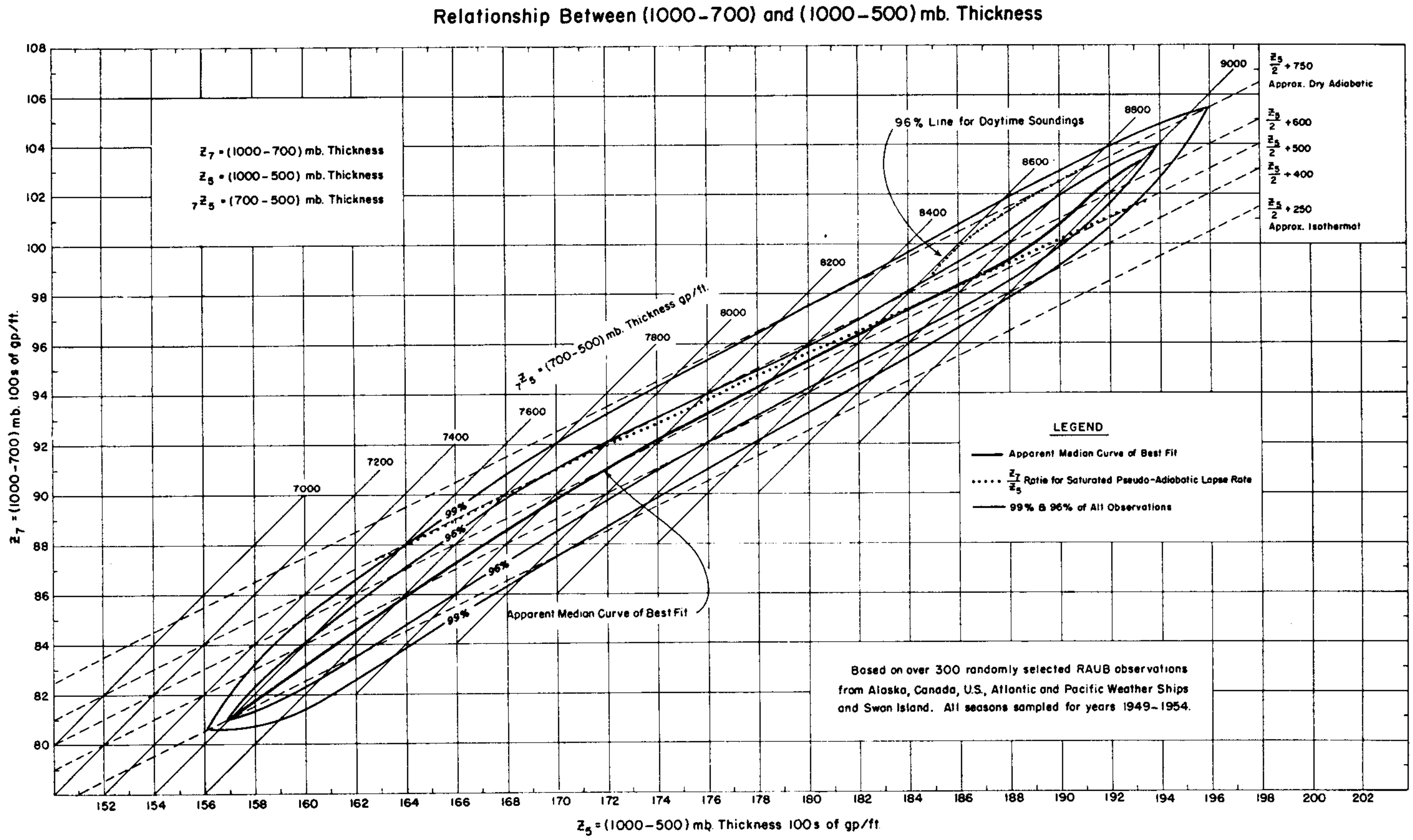

FIGURE 9.—Stability and thickness relationships of two adjacent layers as represented by the graphical solution of the relationship $Z_7=Z_5/2+F$. The spread of the hundreds of observations that entered this relationship is shown by the lines enveloping 99 and 96 percent of the observations.

stability relationship this can be easily computed from individual radiosonde teletypewriter reports.

Similar relationships exist for the $Z_{8.5}-Z_7/2$ or any of the other array of adjacent layers but it must be realized that in each series of thicknesses the median and end values are different. In the $Z_{8.5}-Z_7/2$ relationship the resultant values are all negative with the numbers nearest positive values indicating instability and the numbers farthest from the positive values indicating stability. The adequate mean value in the $Z_{8.5}-Z_7/2$ relationship is approximately -400 gpft., with the extreme values approaching -250 gpft., for which the lapse rate is approximately dry-adiabatic, and -550 gpft., for which the lapse rate is approximately isothermal.

From a few of the lines presented in figure 7 it is easily recognized by comparison with the values previously mentioned that for the greater portion of the region from the Rockies to the West Coast the 1000–500-mb. and 1000–700-mb. thickness relationship indicated unstable air with a similar but smaller area of instability indicated by the 1000–700-mb. and 1000–850-mb. thickness relationship. Stability of the soundings above 500 mb. was not computed.

The distribution of the negative and positive areas, as well as the position of the neutral, or 500 gpft., line, is easily depicted graphically as in figure 9. The curved line, obtained by plotting values of heights at given pressures in a pseudoadiabatic atmosphere for a saturated lapse rate (from [15]), indicates the F values at and above which the airmass is conditionally unstable. In figure 9 this line lies, for the most part, between F values of 500 and 600 gpft. For the $Z_{8.5}-Z_7/2$ relationship, the F values for conditional instability generally range between -400 and -300 gpft. (see table 2). These values must be used in conjunction with a thickness chart and the F chart, for it is at the intersection of the given thickness and F values that the air becomes conditionally unstable.

Even a cursory examination of the charts in figure 7 indicates the relatively good agreement between the two areas of instability, conditional and convective, in the $Z_7-Z_5/2$ patterns. It can be seen that the areas of instability, although located in the vicinity of the west coast on the 1st, became more inclusive of the States of Washington and Oregon on the 2d as moisture increased (table 4) and heavy rains occurred. It was on the 2d that one of the more intense areas of instability, a closed 650-ft. center, appeared south of Burns, Oreg. This relatively large area of instability replaced the stable air mass that on the 1st had extended across central California into Nevada. By 1500 GMT of the 3d (fig. 7C), both

TABLE 2.—*Intersection values for determination of conditional instability*

Z_5	$Z_7-Z_5/2$	Z_7	$Z_{8.5}-Z_7/2$
19,300	500	10,300	−400
19,200	505	10,200	−395
19,100	505	10,100	−390
19,000	510	10,000	−385
18,900	515	9,900	−375
18,800	520	9,800	−370
18,700	525	9,700	−360
18,600	530	9,600	−355
18,500	535	9,500	−345
18,400	540	9,400	−340
18,300	545	9,300	−330
18,200	550	9,200	−325
18,100	555	9,100	−320
18,000	560	9,000	−315
17,900	565	8,900	−310
17,800	570		
17,700	570		
17,600	575		
17,500	575		
17,400	580		
17,300	585		
17,200	590		
17,100	595		
17,000	595		
16,900	595		
16,800	595		
16,400	595		

convective and conditional instability areas had moved eastward from the coastal region but a large area of instability of both types persisted over the Plateau Region and the Rocky Mountain area.

On the lower-level chart showing $Z_{8.5}-Z_7/2$ (fig. 8), conditional instability covered a much smaller area and on the 1st and 2d was confined for the most part to the States of Washington and Oregon. Here again it was inclusive of the region of heavy rain and as the instability area moved north-northeastward on the 3d the attendant precipitation moved with it. It might be interesting at this point to mention that the freezing level on the 1st and 2d was as low as 6,000 feet near the center of the Low and also near the center of this lower-level instability area.

These rather large areas of instability continued over much of the Plateau and the Rocky Mountain regions throughout the remainder of the period although the western limit continued to drift slowly eastward. Numerous thunderstorms or the appearance of distant lightning were recorded during the first six days of the month, and at a few stations more thunderstorms were recorded during those days than was normal for the entire month. Burns, Oreg., was one of the stations with a thunderstorm reported on three separate days during this period. Normally during August only 2 thunderstorms are observed at Burns, Oreg., but during this August there were 7 thunderstorms and 1 night with distant lightning.

Another interesting feature of these charts is the indication of stability over the greater portion of the West by the lower two layers, but instability by the upper two layers. This indicates that most of the thunderstorm activity must have resulted from higher-level instability.

Finally, it might be well to point out another fact that was quite clearly illustrated in the vicinity of Salem, Oreg., on the 3d of the month. On the 2d, as previously mentioned, heavy rain and instability were prevalent over

TABLE 3.—*Precipitation totals and number of thunderstorms August 1–6, 1956*

Station	Precipitation	Normal precipitation	Deviation	Distant lightning	Thunderstorms	Average no. thunderstorms August
WASHINGTON						
Seattle	0.49	0.12	+0.37	0	0	1
Spokane	.38	.06	+.32	1	0	2
Stampede Pass	.12	.26	−.14	0	0	2
Tatoosh Island	.11	.36	−.25	0	0	0
Walla Walla	.40	.03	+.37	0	1	2
Yakima	T	.06	−.06	0	0	1
IDAHO						
Boise	.02	0	+.02	1	1	3
Lewiston	.07	.06	+.01	0	0	3
Pocatello	T	.12	−.12	2	2	7
MONTANA						
Helena	.50	.18	+.32	1	4	7
Kalispell	.26	.21	+.05	1	3	7
Missoula	.25	.18	+.07	1	3	6
OREGON						
Astoria	1.67	.24	+1.43	0	0	0
Burns	.17	.02	+.15	0	3	2
Eugene	.41	.01	+.40	0	1	1
Meacham	.13	.06	+.07	0	1	5
Medford	0	.02	−.02	0	0	1
Pendleton	.36	.01	+.35	3	1	3
Portland	1.06	.06	+1.00	0	0	1
Roseburg	.26	.01	+.25	0	1	1
Salem	.29	.06	+.23	0	1	1
Sexton Summit	0	0	0	0	1	1

the station. However, on the 3d, even though the soundings indicated moisture values at the fixed levels of 850, 700, and 500 mb. as high or higher than those of the 2d (table 4), no rain occurred in this locality as the stability lines had moved to the eastern border of Oregon.

7. PRECIPITATION

Rainfall associated with cold Lows is often variable in amount and intensity and at times practically nonexistent. Nevertheless, experience has shown that usually cut-off Lows in the Washington-Oregon region do produce considerable precipitation especially along the coast and over the Cascades, while inland the totals decrease rather rapidly. Graham [3] has found that 89 percent of the cold Lows that form during the winter months near the Vancouver Island-Washington-Oregon coastline produce precipitation in southern California. And so it would be reasonable to expect considerable precipitation from this type of Low to the north of California during the summer months. Table 3 shows that this storm was no exception to these expectations. The heavier totals did occur along or near the coastal sections of Oregon with Astoria receiving 1.67 inches and Portland 1.06 inches.

It was thought interesting to compare the moisture values of various levels in the upper-air soundings in the vicinity of the cold Low to ascertain the changes that occurred from day to day while it was practically stationary and also changes that occurred later along its path. Therefore the differences between the dry-bulb temperatures and the dewpoint temperatures were obtained for 850-mb., 700-mb., and 500-mb. heights; these values are shown in table 4. The occurrence of moderate or heavy

TABLE 4.—*Dewpoint depression (° C.) at fixed pressure levels, 0300 and 1500* GMT, *August 1–5, 1956.*

Station	Level (mb.)	Aug. 1 03	Aug. 1 15	Aug. 2 03	Aug. 2 15	Aug. 3 03	Aug. 3 15	Aug. 4 03	Aug. 4 15	Aug. 5 03	Aug. 5 15
Tatoosh Island, Wash	850	3	4	5	$\underline{2}$	4	4	2	4	6	5
	700	16	10	3	$\underline{1}$	11	10	5	M	7	10
	500	7	11	8	$\underline{5}$	12	14	10	9	10	14
Seattle, Wash	850	4	1	$\underline{4}$	$\underline{1}$	$\underline{2}$	2	3	4	6	6
	700	17	17	$\underline{5}$	$\underline{2}$	$\underline{2}$	5	7	8	9	15
	500	3	4	$\underline{3}$	$\underline{6}$	(*)	M	15	12	13	14
Salem, Oreg	850	4	2	$\underline{2}$	$\underline{3}$	$\underline{2}$	----	2	4	5	----
	700	8	12	$\underline{3}$	$\underline{5}$	$\underline{1}$	----	7	1	12	----
	500	4	3	10	$\underline{6}$	*M	----	2	5	11	----
Medford, Oreg	850	9	2	4	3	8	6	10	4	6	5
	700	10	6	1	13	9	15	15	5	3	3
	500	M	M	M	10	M	8	5	11	10	7
Spokane, Wash	850	22	22	4	5	8	$\underline{0}$	11	8	10	11
	700	16	15	10	5	3	$\underline{0}$	3	17	16	8
	500	14	5	15	14	*13	*8	M	15	4	10
Great Falls, Mont	850	17	15	10	5	$\underline{5}$	8	15	15	21	12
	700	7	10	10	9	$\underline{7}$	8	10	15	11	7
	500	13	M	2	10	*M	*M	11	12	8	3
Edmonton, Alberta	850	6	9	4	7	7	2	9	$\underline{2}$	$\underline{4}$	8
	700	3	5	10	6	14	0	5	$\underline{2}$	$\underline{2}$	3
	500	4	6	8	15	13	16	M	9	$\underline{3}$	3
Boise, Idaho	850	23	13	15	10	13	15	20	23	24	15
	700	16	14	10	12	11	8	13	14	14	12
	500	M	M	M	M	3	8	11	3	3	M
Ogden, Utah	850	12	9	26	8	24	12	19	23	27	28
	700	5	12	--	18	19	16	15	18	14	21
	500	2	M	8	4	5	11	15	9	3	M

Oakland Calif. Very dry all levels.
Winnemucca, Nev. Very dry most levels.
Lander, Wyo. Generally dry 700 mb. or lower.

Note: Two levels underscored indicates moderate rain within next 12 hours. Three levels underscored indicates heavy rain within 12 hours.
*Depression $<5°$ C. at 600 mb. but dries before reaching 500-mb. height.
M=Motorboating.

TABLE 5.—*Departure of daily maximum and minimum temperature (° F.) from normal, August 2–4, 1956*

Station	August 2 Min.	August 2 Max.	August 3 Min.	August 3 Max.	August 4 Min.	August 4 Max.
Seattle	−4	−21	−8	−7	−5	−2
Spokane	−14	−19	−13	−18	−12	−9
Stampede Pass	−9	−19	−9	−14	−5	−4
Tatoosh Island	−3	−2	−5	−2	−4	−3
Walla Walla	−10	−22	−12	−16	−8	−10
Yakima	−14	−18	−4	−13	−8	−10
Boise	−14	−22	−13	−11	−7	−8
Lewiston	−16	−25	−15	−15	−13	−9
Pocatello	−4	−12	−7	−10	−9	−2
Great Falls	−1	−8	−2	−12	−3	−3
Helena	+4	−3	−3	−8	−8	−4
Kalispell	−10	−16	−16	−20	−24	−11
Missoula	−2	−22	−8	−18	−11	−10
Burns	−7	−22	−14	−13	−3	−8
Meacham	−13	−22	−8	−14	−6	−9
Medford	−11	−18	−12	−15	−7	−10
Pendleton	−9	−21	−9	−15	−1	−11
Portland	−2	−23	−1	−14	−4	−5
Roseburg	−14	−16	−13	−5	−5	−7
Salem	−2	−21	+1	−10	−1	−5
Sexton Summit	−13	−15	−8	−12	−4	−13
Blue Canyon	−14	−13	−11	−10	−13	−12
Eureka	−3	−1	−2	−8	+1	−1
Mt. Shasta	−5	−19	−11	−8	−9	−8
Oakland	−2	0	−2	−7	0	−12
Redbluff	−2	−11	−5	−9	−8	−13
Sacramento	−4	−6	−2	−6	−6	−12
Elko	−13	−16	−17	−10	−12	−7
Ely	−14	−7	−15	−9	−14	−9
Reno	−10	−13	−10	−9	−10	−12
Winnemucca	−7	−15	−21	−8	−9	−5
Salt Lake City	−1	0	−15	−6	−7	−3

precipitation during or within 12 hours following the sounding is indicated in the table by underscoring the values for that sounding. Heavy rain, in this section and

TABLE 6.—*Temperature (° C.) at 5 km., August 1–4, 1956*

Station	August 1 0300	August 1 1500	August 2 0300	August 2 1500	August 3 0300	August 3 1500	August 4 0300	August record from [17]	Years of record in [17]
Boise	−6	−8	−10	−11	−17	−11	−10	−18	7
Ely	−5	−5	−6	−7	−8	−8	−8	−12	7
Great Falls	−8	−9	−8	−7	−8	−9	−12	−14	6
Medford	−10	−16	−20	−17	−11	−9	−11	−15	7
Oakland	−5	−6	−8	−6	−8	−6	−9	−11	10
Ogden	−5	−7	−6	−6	−7	−9	−6	−7	3
Seattle	−16	−17	−19	−17	−18	−16	−14	−22	11
Spokane	−12	−12	−13	−14	−16	−19	−15	−18	12
Tatoosh Island	−17	−18	−16	−14	−15	−17	−13	−15	12

for this season of the year, was defined as a 12-hour total three or more times the normal 24-hour amount. It should be noted that only when the dewpoint depression was near 5° C. or less did moderate to heavy rain occur, and in practically all cases for heavy rain these values had to exist to at least 600 mb. It was also noted that the occurrence of most precipitation was associated with 850-mb. dewpoint values of from +2° to +5° C. In several instances precipitation did not occur even though the moisture content was relatively high at two or three levels. In these cases, it appears that the center of the Low had become removed from this zone of moisture by several hundred miles, and the instability associated with the storm had diminished, thus hindering the production of precipitation. In this regard, attention is directed to figure 7C which clearly defines the area of convective and conditional stability as east of Salem, Oreg., at a time when the air at that station contained high moisture content; precipitation did not occur.

The use of the term "heavy precipitation" to describe the rainfall over most of the States of Washington, Oregon, and western Idaho during the 5-day period August 1–5, is entirely in accord with the class limit values used in the 5-day forecast period. Figure 3B of the preceding article by Andrews [16] clearly indicates that this rain made an important contribution to the total precipitation for August, for there were only two rainy periods during August with both of about equal intensity at many of the stations.

8. TEMPERATURE DEVIATIONS

That both the maximum and minimum temperatures reported by the first order stations during the period of August 1 to 5, 1956, were much below normal is shown in table 5. In this table the entire period was not included but a sufficient interval was indicated to note the scope of the temperature deviations. Lewiston, Idaho reported the greatest departure of maximum temperature in the period on August 2, a reading which was 25° F. below the normal high for that date. But it will be noticed that numerous other departures in excess of 19° F. were observed. The greatest anomaly in relation to minimum temperatures was recorded at Kalispell, Mont. where the minimum of August 4 was 24° F. below normal.

Another indication as to the intensity of the cold air associated with the cold pool is presented in table 6 which lists the 5-km. temperature observed August 1–4 and the extreme minimum 5-km. values for August as reported in [17]. It is realized that the minimum values presented in [17] are, in many cases, from short-period records, and in only a few instances do they represent a decennium. However, it will be noted in table 6 that Medford, Oreg. at 0300 GMT August 2 had a temperature of −20° C. at 5 km. which was 5° C. below the extreme value (period: September 1939 to December 1945). At Oakland, Calif., well to the south of the cold pool center, the temperature at 0300 GMT of the 4th was within 2° of the low for August obtained in a period of record from September 1936 to December 1945. Spokane, Wash. reported, at the same level, a reading of 1° C. below the low that was established during the period from July 1934 to December 1945.

A summation of the departures from normal of temperatures that occurred during the interval of August 1–5, is presented in figure 10 which clearly depicts the fact that the region had much below [1] the normal average temperature during that period. Absolute values are not indicated but the zone encompassed by the two shaded areas experienced an accumulated total of 25° F. or more below the daily normal for those 5 days. The area of inner shading represents an aggregate deviation of 50° F. or more during the same 5 days. The station with the greatest deviation sum during the 5 days was Lewiston, Idaho with a total of 71° F. indicating that during August 1–5 the maximum and minimum readings at Lewiston averaged approximately 15° below normal each day. The temperature anomaly for August (see Andrews, fig. 2B, p. 307 of this issue) indicates that the Western States recovered from these low temperatures to a considerable extent prior to the end of the month.

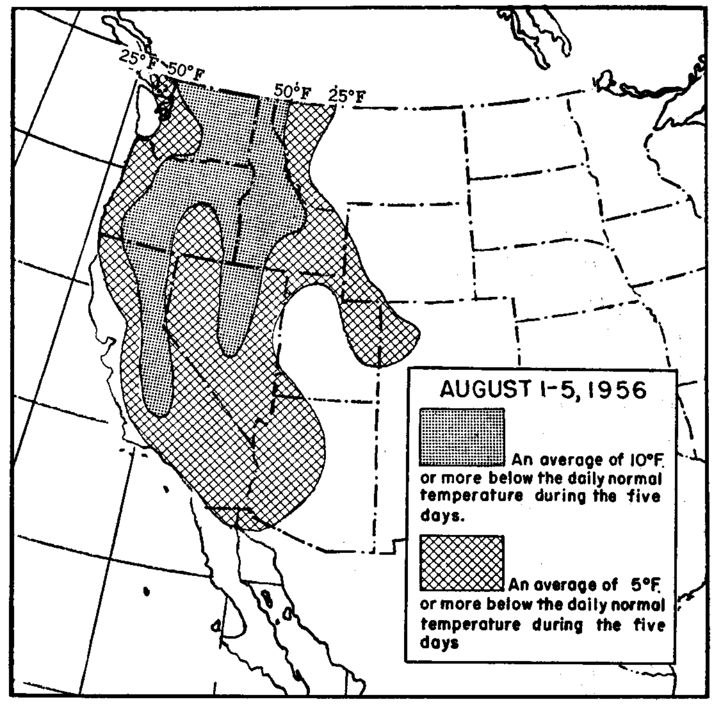

FIGURE 10.—Total temperature deviation from normal in ° F. for the 5-day period August 1–5, 1956. The entire shaded area averaged 5° F. or more per day below the normal August daily average temperature during this 5-day period. (5° F. was used since this is the approximate class limit value used by Extended Forecast Section in this area during August to represent much below normal temperatures.) The inner shaded area averaged 10° F. or more per day below the normal daily average with the maximum departure approaching −15° F. per day at Lewiston, Idaho.

9. SURFACE TEMPERATURES AND THICKNESS RELATIONSHIPS

It is generally believed that a definite relationship exists between the surface temperatures and the thickness of the overlying layer, and indeed under certain conditions the correlation between areas of anomalous thickness and areas of anomalous maximum or minimum temperatures appears to be quite good. Because the departures from normal maximum and minimum temperatures were extremely large during several days of the period of this study it was thought that a presentation of the temperature relationship to the departure from normal thickness as obtained from the NWAC 30-hour 1000-500-mb. prognostic thickness chart would be informative. The actual observed isopleths of thickness departures from normal are not presented in this study but the area and intensity of the departures from normal 500-mb. height (fig. 2B to 6B) were in close propinquity with the thickness departures from normal and may be used for comparative purposes.

In a case study on this subject, Kibler, Lennahan, and Martin [18], and in later research Allen and Ellis [19], found that this relationship is somewhat dependent upon the physical and geographical location of the station as well as the season of the year. For example, large water bodies tend to influence the temperature values by their moderating effect. There was some indication that during periods of large thickness and high temperature, deviations from the regression line tended to be greater than under conditions of low thickness and low temperature.

A higher correlation between thickness and surface temperature might be expected when there is good air movement and also when the ground covering remains uniform over the season for which the relationship is computed.

There are many other factors that enter into this problem of forecasting the maximum and minimum temperature from the deviation of thickness from normal. One of the most troublesome of these is the fact that the maximum or minimum temperature may occur at any time during the 24 hours, possibly as much as 12 to 18 hours prior to or after the normal time of maxima or minima. Thus an individual maximum or minimum temperature may not be related to the current airmass over the station

[1] This value, approximately 5° F. per day departure, was ascertained as the value used by the Extended Forecast Section in that region during the month of August for their class limits of "much below normal".

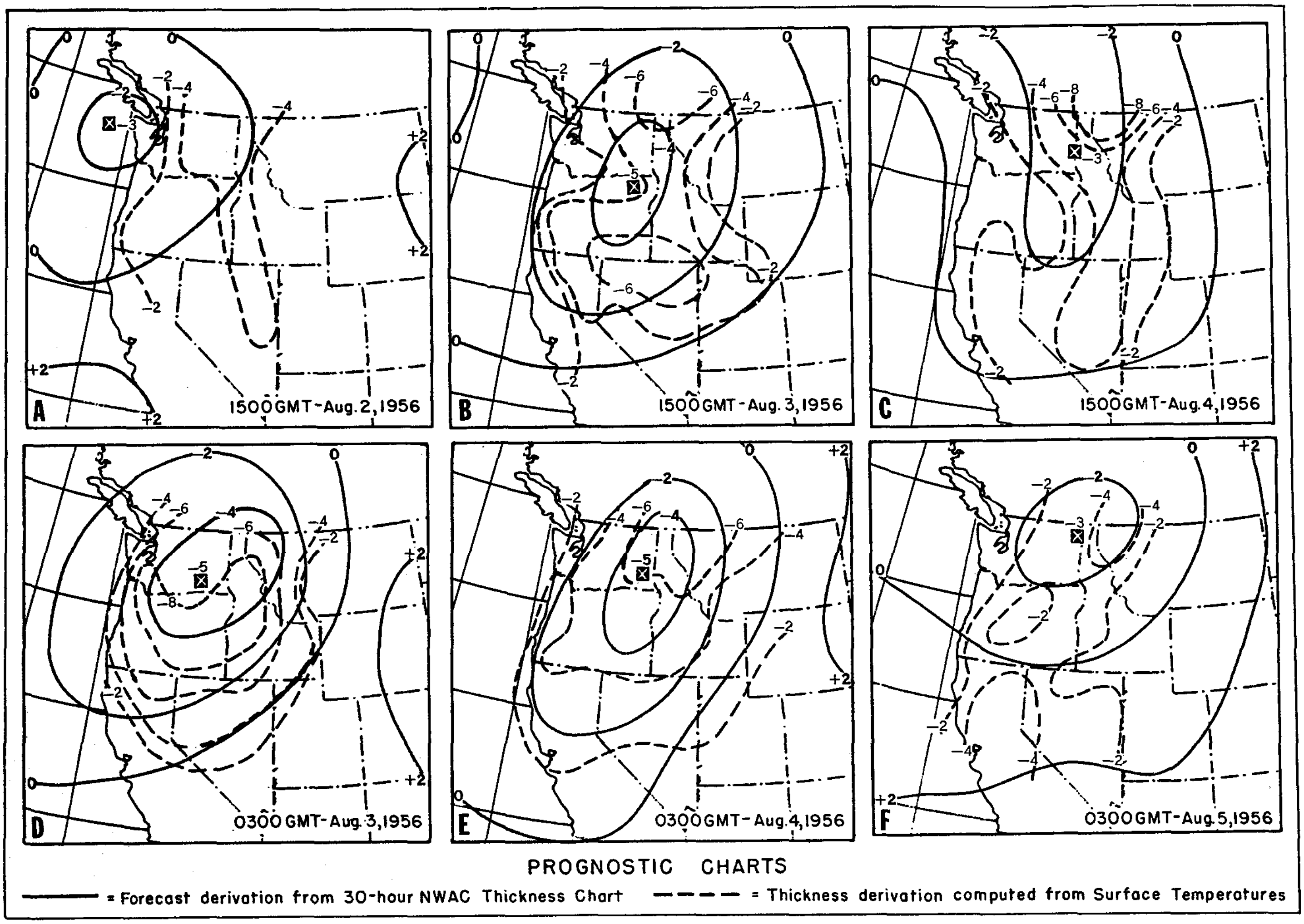

FIGURE 11.—Composite charts consisting of the 30-hour prognostic 1000-500-mb. (Z_5) thickness anomalies (solid lines) as derived from the Z_5 thickness prognoses and the normal thickness chart for August, and the actual deviations of temperature maxima or minima from normal converted into quasi-thickness isopleths (dashed lines). This furnishes a picture of the relationship between the deviations of temperature and deviations of thickness. Actual thickness values are not presented but as previously stated are similar in area covered and value to the departure from normal height in this study. Generally in these charts the relationships of departures of temperatures to Z_5 departures are in quite good agreement as to intensity and area covered. (A) 1500 GMT, August 2. (B) 1500 GMT, August 3. (C) 1500 GMT, August 4. (D) 0300 GMT, August 3, (E) 0300 GMT, August 4. (F) 0300 GMT, August 5, 1956. Parts A, B, and C employ minimum temperatures; parts D, E, and F, maximum temperatures.

at the time of the prognostic chart but to a later outbreak of cold air or influx of warm air. Hence the correlation may be improved if a concurrent surface temperature is used instead of the maximum or minimum. In a 36-hour prognostic thickness chart, the indicated departure from normal would be applicable in the forecasting of maxima or minima at times a short period in advance, but future modification would have to be employed for the forecast to be extended beyond that time.

Minimum and maximum temperatures are also frequently affected by local cloud cover, drainage winds, and other orographic factors, hence a generally applicable relationship between the departures from normal thickness and anomalous daily temperature extremes will not likely be found. However, over much of the country, the surface temperature extremes are influenced by the mean temperature of the overlying airmass to a greater extent than by any other factor, and the areas of thickness departure from normal provide a definite guide in forecasting temperature anomalies. Some of the problems with regard to large positive anomalies of thickness were discussed by McQueen and Shellum [11].

Charts illustrating the departure of the prognostic 1000-500-mb. thickness from normal as prepared twice daily by NWAC are shown by the solid lines in figure 11. In figure 11, parts A, B, and C, are for the approximate time of the minimum temperature and parts D, E, F, are for the time of the maximum for August 2, 3, and 4. It will be noted that the dashed lines represent pseudo-thickness lines obtained by converting the anomalies of maximum and minimum temperatures at each first order station into thickness deviation lines by using 5.4° F. as

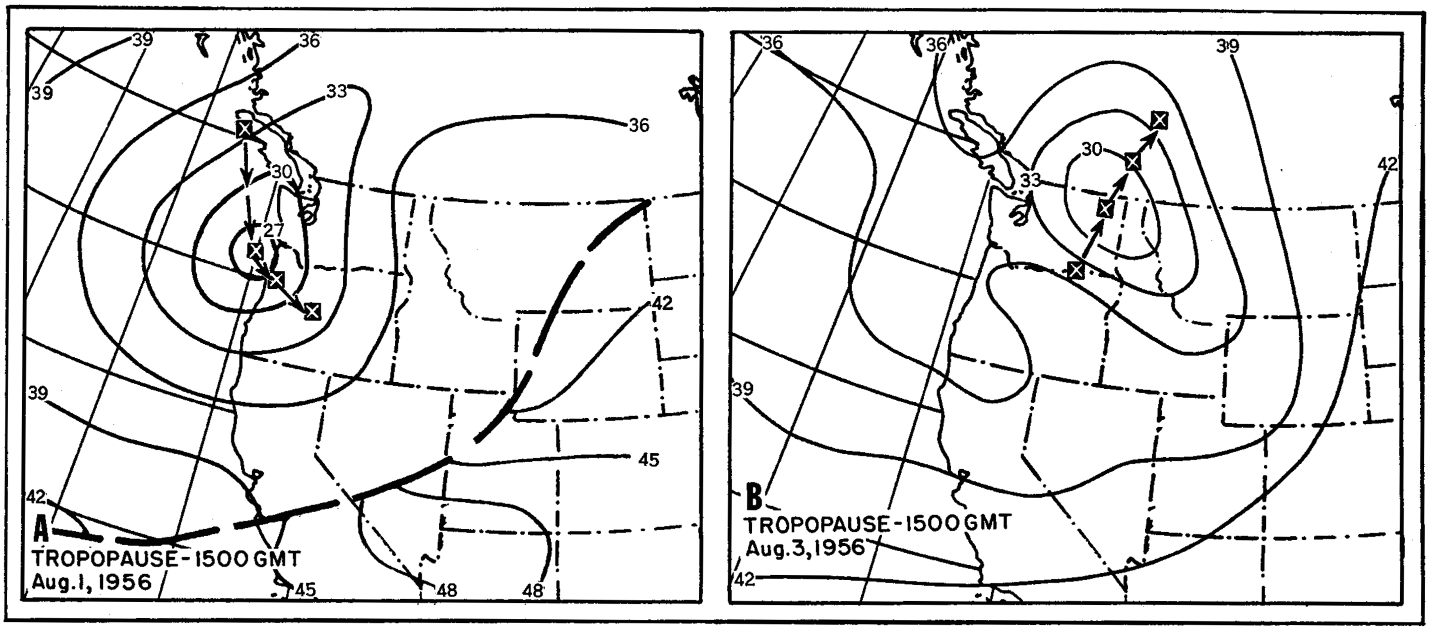

FIGURE 12.—Tropopause height charts over the Western States (labeled in thousands of feet) with track of 12-hour positions of tropopause minimum. (A) 1500 GMT, August 1, 1956. Heavy line is tropopause break line. This chart clearly indicates the depth and intensity of the cold air which lowered the height of the tropopause to 26,000 feet at the center on this date. (B) 1500 GMT August 3, 1956. Note the persistency of this cold pool and the little modification that ensued. This pool of cold air was easily followed for the next several days as it drifted northward and eastward across Canada.

the relationship for each 200 ft. of thickness, since mean temperature changes of this magnitude are equivalent to 200-ft. changes in 1000-500-mb. thicknesses.

It is readily apparent that the forecast thickness anomaly and the observed pseudo-thickness in nearly all cases were in close relationship to each other. It may also be seen, by use of the height anomalies as a fairly good approximation of the thickness anomalies in this study, that the pseudo-thickness lines obtained from the temperatures are in contiguity with the height anomalies.

10. TROPOPAUSE

Because the center of the cold Low was poorly defined from the surface analysis it was thought that its definition on the NWAC tropopause height charts would be of interest. Two of these charts at 48-hour intervals are reproduced in figure 12. They clearly indicate that the cold pool was well-defined in the upper levels of the atmosphere with the tropopause surface near 26,000 ft. at 1500 GMT of August 1, 1956. This may be among the lowest tropopauses to have occurred in that region during the month of August, although comparative data are not available for checking this possibility. The 5-km. temperatures were near or at record low values during this time and there appeared to be but minor changes of lapse rate between that level and the 8-km. level which was the approximate height of the tropopause on August 1.

During the period under study the tropopause minimum remained about vertically above and below the low centers on the constant pressure charts and its sequence of motion was in agreement with the motion of the low centers at other levels. After 1500 GMT of August 3, there was gradual filling of the cold pool with attendent temperature modification; for example, the temperature of −50° C., at 26,000 ft. on August 1, was associated with heights of about 35,000 ft. by August 5.

ACKNOWLEDGMENTS

The writers wish to express their appreciation to the staff members of NWAC for helpful suggestions and the reviewing of the article, and to the Daily Map Unit of the Weather Bureau for detailed drafting of the figures.

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